CHAPTER II. SEA ICE DYNAMICS

SECTION 1. GEOSTROPHIC WIND FIELDS

This section describes what is meant by the term geostrophic wind, which is the horizontal wind velocity for which the coriolis acceleration exactly balances the horizontal pressure force. This term reflects what is probably the most fundamental concept in meteorology. Although we will not be able to develop our understanding along lines of mathematical physics, we will be able to compile a working knowledge of this term in a way which is useful to the ice observer and short-term forecaster.

1.1 ATMOSPHERIC PRESSURE

The basic tool in determining the distribution of wind velocities over a region of the earth is the barometric pressure chart. This chart is a contour map of atmospheric pressure at some prescribed altitude. The terminology used requires some explanation.

The millibar is the basic unit of atmospheric pressure measurement. By definition, this is 1/1000th of a bar which is the pressure generated by a force of 100,000 newtons acting on an area of one square meter. Long-term measurements have shown that the average sea level atmospheric pressure is 1,013.25 millibars (abbreviated mb). In terms perhaps more familiar, this is 14.7 pounds per square inch. The atmospheric pressure record used most often in sea ice analysis is called the Synoptic Mean Sea Level Pressure Analysis. These charts display contours of atmospheric pressure corrected to mean sea level. Other charts sometimes used give altitude contours for the height of a particular atmospheric pressure above mean sea level (for instance, the 750 mb chart). At 500 mb, half the atmosphere is below and half above the point of observation. This turns out to be a height of around 5 km or 3 miles. Obviously, the atmospheric density does not vary linearly with altitude, as a pressure of 1 mb is not reached until heights on the order of 30 km are attained.

Atmospheric pressure is measured by instruments called barometers. It is quite easy to make very sensitive barometers. An altimeter is little more than a sensitive barometer, which to be accurate, must be calibrated against the sea level pressure directly beneath its location. In order to measure pressures at various altitudes, balloons are released which climb at a known rate and radio back pressures measured by a "throw-away" barometer.

1.2 MOVEMENT OF AIR RESULTING FROM ATMOSPHERIC PRESSURE SYSTEMS

The variations in atmospheric pressure are complex and arise from a number of factors. The sun is, of course, the primary source of energy to the earth's atmosphere. This energy is absorbed in some places, like oceans, and reflected in other places, such as areas covered with snow and ice or clouds. A great deal of this energy finds its way into the atmosphere, both in the form of direct heating and the added heat of evaporated water. As a result of the complexity of this process, regions of relatively high and relatively low atmospheric pressure are created. Much of what we commonly call "weather" is created as these pressure systems attempt to even out and come to equilibrium. As we shall see, *this process would be much simpler if the earth were not turning one* revolution per day on its axis. Were the earth not rotating, areas of high and low pressure would come to equilibrium by simply exchanging sufficient air to possess equal pressures. The rate of exchange of air or wind speed would depend on the spatial rate of change of pressure, or pressure gradient. Hence, the greater the pressure change in mb per unit distance, the greater the wind. Furthermore, the wind would be along the direction of greatest change in pressure (i.e. perpendicular to the contours of equal pressure or isobars).

If the earth were not rotating, the acceleration experienced by the air particles would be given by:

$$a_p = -\frac{1}{\rho}\frac{\Delta p}{\Delta x}$$

where: ρ is the atmospheric density

$\Delta p/\Delta x$ is the change in pressure over a given distance divided by that distance (this is the pressure gradient). The negative sign arises because the air is accelerated from high pressure to low pressure, in the opposite direction to the positive pressure gradient.

The formula reflecting the affect of the earth's rotation on wind speed is given in II-1.2.2.

1.2.1 Coriolis Force. When we fire a bullet, we expect that in the absence of winds or other external forces, it will travel in a straight line (although it will change its elevation as it travels along this straight line). But where is this straight line actually drawn? About three centuries ago, Isaac Newton came to the conclusion that in fact the straight line is drawn with respect to the distant stars and not the surface of the earth, which is actually turning on its axis. Sometime later, a man named Coriolis determined the exact value of this small discrepancy.

In order to understand this concept, imagine standing exactly at the north pole on a day in early spring when the sun is just above the horizon (where it will stay all day as the earth rotates). Imagine that you have placed a target 300 m away and just as the sun (a reasonably distant star) passes behind the target, you fire a bullet with a muzzle velocity of 300 m/sec at the target.

In reality, the bullet is traveling toward the sun and the target just happens to be between you and the sun when the bullet was fired. It will be 24 hours until the target is between you and sun again. Meanwhile, the target is moving eastward (counterclockwise) at a rate of 2.2 cm per second. (The target travels 2π x 300 m in 24 hours or 1,884 m. This divided by [24 hours x 60 min/hour x 60 sec/min] or 86,400 equals .022 m/sec or 2.2 cm/sec.) Therefore, to actually hit the center of the target, the bullet should have been aimed 2.2 cm to the east of the target's center when it was fired.

Coriolis realized that most of us here on the earth think of our reference frame as the earth and not the fixed stars. Not being aware of the fixed star reference frame, most of us would fire the bullet straight at the target. We now know that it would land 2.2 cm to the west of the target's center, as if a mysterious force actually acted on the bullet to accelerate it off target. All moving things on the earth's surface are subject to coriolis acceleration. However, usually the deflection is not noticeable.

Coriolis worked out the equation to describe this acceleration in the horizontal plane. It is given by:

$$a_c = + 2w\, v \sin \phi$$

where: w is the rotational velocity of the earth in radians per sec.
v is the velocity of the object
ϕ is the latitude of the object

The reason for the dependence on latitude may be a little puzzling at first, but a little thought will explain why. Consider the above target practice being carried out at the equator instead of the north pole. If the bullet were fired at a target due north or south, it would travel parallel to the earth's axis (which is fixed with respect to the stars) *and would not be deflected. Similarly, if the target were due* east or west, the bullet would be deflected up or down but not north or south. There would be no deflection in the horizontal plane.

1.2.2 Geostrophic Winds. What really happens to the particles of air in our atmosphere is the result of the interaction of the pressure gradient acceleration and the coriolis acceleration. A complete explanation of how these factors combine is beyond the scope of this handbook. In fact, even an accurate explanation is beyond the scope of many meteorological texts. However, the result is a steady wind along the pressure isobars given by:

$$v_g = \frac{1}{2w \sin \Phi} \cdot \frac{1}{\rho} \frac{\Delta p}{\Delta x}$$

Here we recognize the first term as related to the coriolis acceleration and the second as the pressure gradient acceleration.

As a result, instead of atmospheric high and low pressure cells creating winds directly across the isobars, the winds created are along the isobars. The direction along the isobars is easy to understand. Imagine looking down on a high pressure cell in the northern hemisphere. Because of the high pressure, the air wants to expand in all directions. Consider air particles attempting to move southward in terms of the example of the southward-fired bullet. Coriolis acceleration will direct the air to the west. Similarly, air particles attempting to move eastward will be deflected toward the south. The result of this deflection will be a clockwise motion about the center of the high pressure cell. Similarly, coriolis acceleration will produce counterclockwise motion about a low pressure cell. A little thought will show that these motions are reversed in the southern hemisphere.

The geostophic wind related above describes to a first approximation the motion of air in our atmosphere. However, it does not take into account a number of factors such as the motion of pressure cells and increasing or decreasing pressure. Also, surface effects can completely distort geostrophic winds so that the geostrophic wind concept can only safely be applied to altitudes somewhat higher than the geographic relief in an area.

1.2.3 Modification of Geostrophic Winds by Surface Effects. The winds of most interest to sea ice forecasters are those at the surface of the ice. Hence, the "surface" chart is used. However, at this level, the motion of the air is not purely geostrophic: (1) Even for a smooth surface there will be frictional effects. (2) Mountains and hills can considerably alter surface winds. (3) Cold air generated on large glacial ice masses can create violent local winds. We shall consider these in turn.

Before discussing these modifications to surface winds due to local causes, it is useful to consider the manner in which pressure charts are created. Barometric pressure is only measured at a relatively few locations, particularly in the arctic and antarctic. Yet, detailed pressure charts are drawn based on these data. This is much like constructing a geographic contour map based on a handful of altitude measurements. The result is an overall pattern which is generally correct, but the precise configuration of the isobars cannot be obtained. Therefore, the geostophic winds, which are determined by the density of isobars ($\frac{\Delta p}{\Delta x}$ is simply the number of isobars per unit distance), are also only generally correct. There is simply no way to account for local effects except through observation and experience.

1.2.3.1 Frictional Effects. The standard height for surface wind measurements is 10 m. This is usually sufficiently high to eliminate purely site-specific effects. Yet, even above smooth ice, the winds at 30 m are not geostrophic - that is, along the isobars. In fact, a height of around 500 m is required to reach geostrophic winds. Below that altitude there is sufficient friction to appreciably slow the wind.

This slowing alters the balance between the pressure gradient acceleration which remains largely constant, and the coriolis acceleration. (Remember, coriolis acceleration is directly proportional to v.) As a result, as the surface is approached, the winds are deflected less by coriolis acceleration and have a component along the pressure gradient. The amount of this effect is around 15° over open water and around 20-25° over ice.

1.2.3.2 Mountains and Hills represent barriers to air motion and can alter completely the direction of winds in their vicinity. In some of these locations, the local winds show a strong tendency to be along a particular direction regardless of the geostrophic pattern. Local wind patterns such as these can considerably alter sea ice motion for relatively large distances (20-50 km) from the mountain barriers. These effects are only known through site-specific experience. These alterations are known as orographic effects.

1.2.3.3 Offshore Winds are also caused by katabatic winds. These are winds created in sloping mountain valleys as the air is cooled in the evening and, being heavy, "slides" downslope. This effect can be particularly strong from glaciers and sloping glacial ice fields. Antarctica is particularly well known for katabatic winds. When these winds reach the water, they can move ice floes considerably seaward, creating an appreciable flaw. The effect of katabatic winds is limited to distances less than 40 km from coasts.

SECTION 2. OCEANIC CURRENTS AND TIDES

In addition to winds, the motion of the ocean has a major influence on the motion of ice. There are several sources of oceanic currents causing them to be quite complex at times. The motion of water due to tides can also be complex, particularly in shallow and long embayments. This short section cannot possibly attempt a complete treatment of this subject. However, some topics of interest to the ice analyst can be outlined.

2.1. OCEANIC CURRENT SYSTEMS

Several centuries of observations have yielded considerable knowledge of the major current systems in the earth's oceans. Yet, as late as 1961 a major and important current in the equatorial Pacific Ocean was discovered. Although published charts of oceanic currents appear to imply steady and constant currents, this is not really the case. These charts represent averages and generalizations. The two accompanying charts show the known current systems in the arctic and antarctic. Where data are available, the magnitude of these currents is given in knots as well as the direction of the current (Figures II-1, and II-2).

The magnitudes of oceanic currents are not large compared to wind velocities. Nevertheless, they are much more constant and, therefore, can add a significant component to ice drift, especially over periods of time greater than a day.

In addition to their importance as a means of transporting ice, currents are a major source of energy transport to arctic regions. This energy is supplied in the form of storms generated from the warm currents as well as the transport of warm water. In later sections, the impact of current systems will be discussed in terms of ice behavior in specific regions.

2.2 OCEANIC TIDES

The detailed description of tidal behavior and related causes can easily fill an entire book. All we can hope to do here is present the sea ice analyst with sufficient background to understand tidal phenomena which might have a bearing on sea ice behavior.

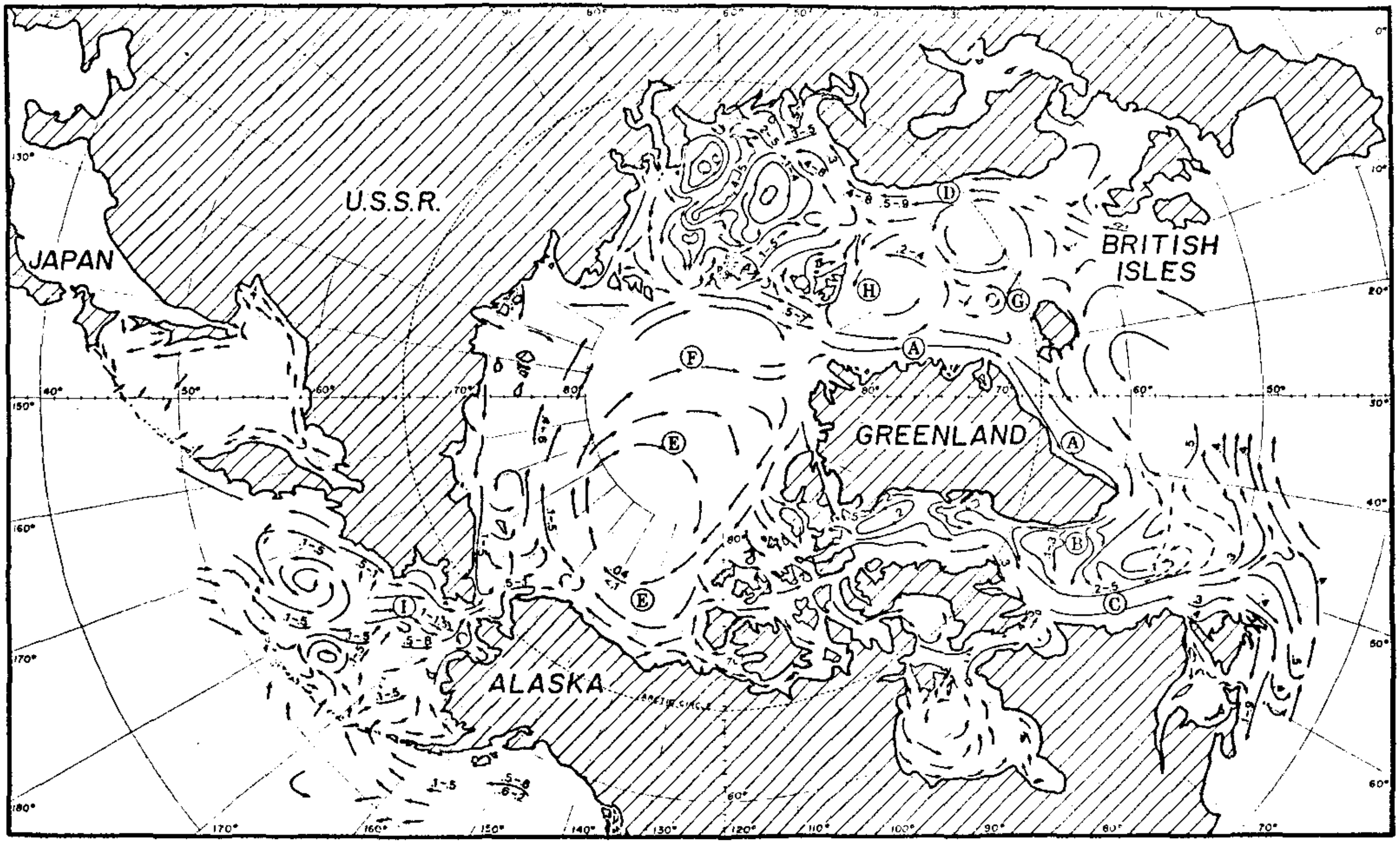

Figure II-1. Arctic Oceanic Currents. Current values shown are in knots. Current names are as follows: (A) East Greenland Current, (B) West Greenland Current, (C) Labrador Current, (D) Norwegian Current, (E) Beaufort Gyre, (F) Trans-polar Drift Stream, (G) Jan Mayen Current, (H) West Spitsbergen Current, (I) Pacific Current.

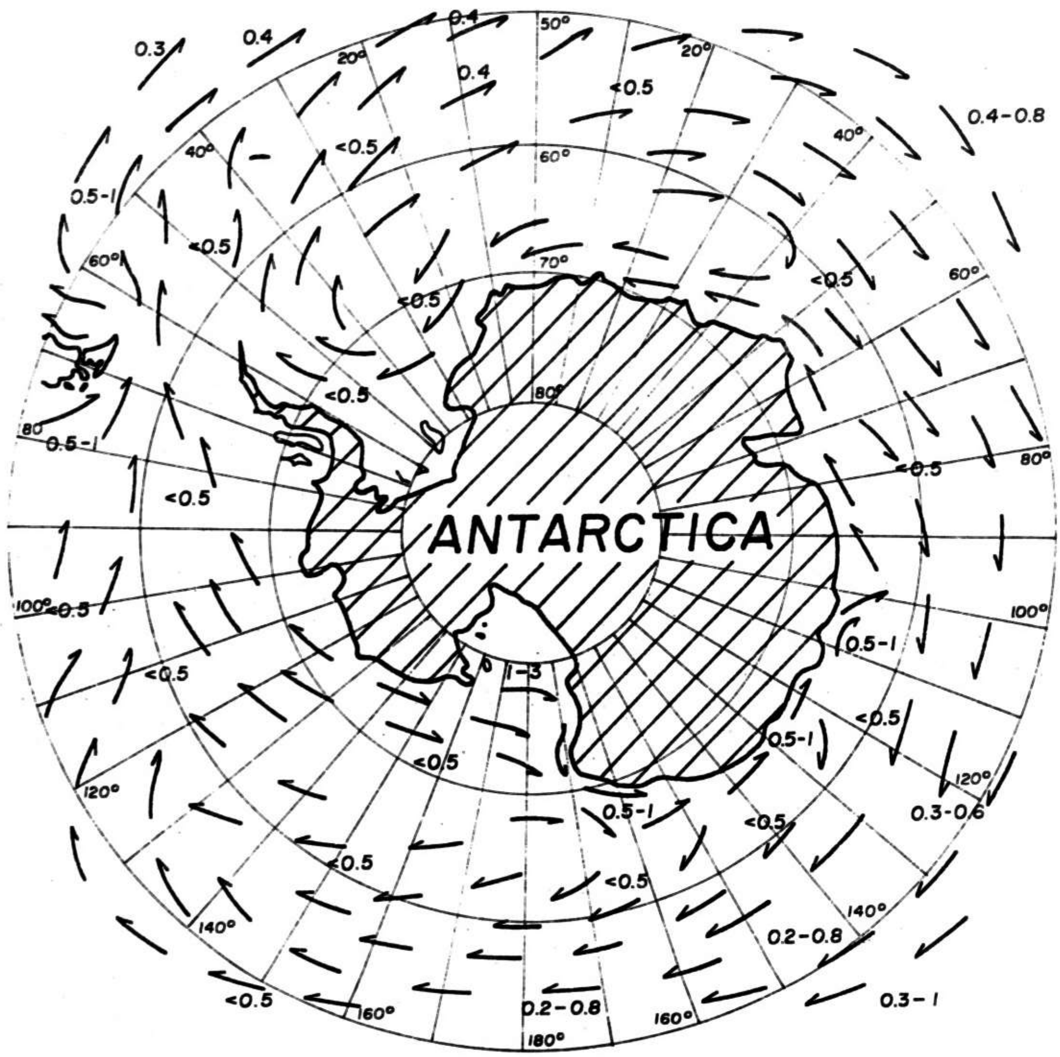

Figure II-2. Antarctic Oceanic Currents.
Current values shown are in knots.

2.2.1 Causes of Tides. Oceanic tides are not only caused by gravitational attraction of the moon and sun for the earth's water but also by the centripetal acceleration of the planet's water surface as both the earth-moon pair and the earth-sun pair revolve around their common center of mass.

The effect of either the sun or moon acting alone would be to cause two high tides and two low tides per day. On a planet covered completely with a deep ocean, the high tides would occur when the celestial body was directly overhead and directly underneath the point of observation. However, the tides caused by the sun and moon add to yield two high tides and two low tides simply as addition of the two separate tidal waves. However, the timing of tides changes during the month because the relative position of the moon and sun changes over that period. Twice a month, when the sun and moon are together and when they are opposite each other, the tidal waves of each combine to cause very high tides (spring tides). Similarly, when the sun and moon are 90° apart (first quarter and third quarter), the tides are their smallest (neap tides).

Even during spring tides, one cannot expect the high tides to occur at noon and midnight. This is because the actual configuration of the ocean and shoreline embayments alters the timing of tides considerably. (Although the relative positions of the tide states will remain in phase.) Because of these site-specific effects, a tide table must be consulted for tide stages at various locations.

2.2.2 Magnitudes. In the open ocean, tidal magnitudes are only a few centimeters. The really noticeable tidal effects are caused by interaction of the tidal bulge with the shore where water can pile up in shallow areas. The coastal tidal range can vary from ten or twenty cm on the shore of the Beaufort Sea to nearly 16 m in the Bay of Fundy, Nova Scotia. Again, tide tables must be consulted to find the tidal range to be expected at various locations.

2.2.3 Geometric Effects. The particular geometry of a sound, inlet, bay or any partially closed body of water will have a great bearing on the tidal range and timing within that feature. In addition, because coriolis acceleration causes a turning motion to the right in the northern hemisphere [and to the left in the southern hemisphere (sh)], tidal waters entering a partially enclosed body will tend to pile on the right ($left_{sh}$) shore and exiting will tend to accumulate on the left ($right_{sh}$) shore. (Right and left taken in terms of facing the entrance of the water body from the outside.) Furthermore, the tidal range will be much larger on the right ($left_{sh}$) shore.

In some waterways the speed of advance of the incoming tide can be several knots and in extreme cases, a tidal bore or wall of advancing water will be created. Turnagain Arm of Cook Inlet, Alaska can be host to both sea ice and a tidal bore.

2.2.4 Resultant Ocean Particle Movement. Here it is instructive to consider the actual movement of water due to tidal currents. (The motion of ice on a moving body of water will be discussed later.) In a narrow waterway, the tidal currents are greatest when the tide state is changing most rapidly. This occurs roughly halfway between high and low tides. The current which preceeds high tide is called the flood tide current and the current which preceeds low tide is called the ebb tide current. In this case, the path of a particular particle of water is roughly one-dimensional. However, in the case of a larger water body, a particular particle of water tends to undertake an elliptical path as a result of the coriolis acceleration described above. These ellipses tend to be counterclockwise in the northern hemisphere and clockwise in the southern hemisphere (viewed from above the earth). When considering the effect of tides on ocean particle movement, it is important to remember that the tidal ellipse is superimposed on any other current which may be acting in the area. In that case, the particle motion will not necessarily be a closed figure.

2.3 PERTURBATIONS TO OCEANIC FLOW ARISING FROM METEOROLOGICAL CONDITIONS

Meteorological conditions can alter oceanic flow by means of two major mechanisms.

2.3.1 Wind Stress on the water surface obviously creates waves. More importantly for this discussion, it creates a surface current. The precise relationship between wind and the induced currents is not well known. However, there is general agreement that the force on the water surface (the wind stress) is roughly proportional to the wind speed squared. Hence, a 40 knot wind will produce 16 times the water stress that of a 10 knot wind. Because of coriolis acceleration (described in 1.2.1 of this chapter), the water (and ice moving freely upon it) will actually move to the right ($left_{sh}$) by 20 to 40 degrees. Wind stress can cause a storm surge tide as great or even greater than astronomical tides. In the Beaufort Sea, for instance, the astronomical tide is less than one meter. In that same region storm surges of several meters have deposited ice high above the normal high tide line.

2.3.2 Sea Level Tilt can be created by the piling of water by wind stress. This can become particularly important when a storm surge encounters land and the piled water must flow away according to its slope just as a river runs downhill. Again, coriolis acceleration becomes important to the water moving downslope, tending to turn it to the right in the northern hemisphere.

Sea level tilt can also result from atmospheric pressure variations. This mechanism has been found to be an important factor in creating a southward flow of water (and ice) through the Bering Strait between Alaska and Siberia. Barometric pressure-induced sea level tilt is particularly effective in shallow and constricted waters such as the Bering and Chukchi Seas.

SECTION 3. MOTION OF ICE UNDER THE COMBINED EFFECTS OF WINDS, CURRENTS, AND CORIOLIS ACCELERATION

3.1 HISTORICAL VIEW

In 1893 a Norwegian, Fridtjof Nansen, initiated a transpolar drift across the Arctic Ocean aboard a specially constructed ship, the Fram. Among the many observations made by this expedition was that ice floes on the water surface drifted between 20 and 40 degrees to the right of the surface wind direction. Nansen correctly recognized this as an effect of coriolis acceleration and suggested to a Norwegian physicist, Ekman, that this effect might warrant theoretical investigation. Ekman's subsequent work resulted in the accurate prediction of a large range of oceanic effects related to coriolis acceleration and other phenomena.

The U.S.S.R. has long depended on her northern ports and seaways for transportation of goods within that country and for export/import purposes. Since the late 1920's, the soviets have sent numerous expeditions to the arctic to conduct detailed scientific investigations. By 1943, sufficient information had been obtained that N.N. Zubov could write a massive book of 500 pages (when translated into English) at a time when very little scientific work on arctic ice had been performed in other countries, particularly the United States. Zubov's work touched on almost every imaginable subject related to oceanic ice. Of interest to us here will be his treatment of ice drift as a result of synoptic winds.

Since that time, the soviets have continued their efforts and considerable work has been done by other countries as well. In recent years, the U.S. effort has had two main directions. The first was a large experiment and theoretical program to understand and develop a numerical model to describe drift ice behavior under bounded conditions such as exist in the U.S. portion of the Beaufort Sea. This project was referred to as AIDJEX (Arctic Ice Dynamics Joint Experiment). The second aspect, a major part of which is underway at this writing, is to understand and model the behavior of drift ice in the marginal ice zones (MIZEX). In these regions, ice is growing or melting and usually freely drifting under the direct influences of winds and currents. It is generally presumed that this contrasts with the behavior of drift ice under the bounded conditions studied under AIDJEX where forces can be transmitted considerable distances (on the order of hundreds of km) within the drift ice. However, the importance of this contribution to the movement of the ice is a matter of debate even today. Fortunately, it does not appear to be of great importance to predicting the location of the drift ice edge.

The movement of sea ice resulting from winds, currents, and coriolis force is a complex subject. A great deal of effort has been made to develop sets of equations which allow ice analysts to make predictions of ice motions based on geostrophic winds (taken from pressure charts) and currents. There are many difficulties. First, there is the question of whether the ice in question is open drift ice, closed pack ice, a few isolated floes, or the drift ice edge. Second, it is often difficult to obtain adequate knowledge of the geostrophic winds in polar regions because of inadequate barometric measurements. Finally, currents are hard to measure and in some areas can change on a relatively short time scale, making the use of standardized current charts questionable.

The problems mentioned above present difficulties not only to the sea ice analyst, but also to the scientists attempting to determine the equations describing ice motion. Despite these problems, some rather consistent equations have been derived.

3.2 OVERVIEW OF THE RELATIONSHIP BETWEEN GEOSTROPHIC WINDS AND ICE MOVEMENT WITHIN THE DRIFT ICE

3.2.1 Geostrophic Winds and Direction of Floe Motion. Although winds at sea level provide the force for moving ice, surface wind values are generally not available to the ice analyst. In order to derive surface wind values, geostrophic wind values must be altered by an appropriate angle to correct for surface friction as described in II-1.2.3.1. This angle is around 25° to the left. However, as described in II-1.2.1, coriolis force alters the direction of moving objects on the earth's surface, turning them to the right (in the northern hemisphere). For sea ice, this angle ranges from 30° to 40° from the surface wind. As a result, sea ice motion is usually between 5° and 15° to the right of the geostrophic wind.

3.2.2 Geostrophic Winds and Ice Velocities. Fluids generally exert square-law forces on objects placed within them. That is:

(1) $F = K\ V^2$

or the force is equal to the velocity squared times an empirical constant which is related to the viscosity of the fluid and surface friction characteristics of the object. Sea ice has two such forces acting upon it: winds and currents.

Although we do not know the ratio between the geostrophic winds and the surface winds, we can assume it is a constant factor, K_1, so that:

(2) $V_S = K_1\ V_G$ where V_S = surface wind speed and
V_G = geostrophic wind speed

Using equation (1), the propelling force on an ice floe due to surface winds is:

(3) $F_P = K_2\ V_S^2$ where K_2 is a constant combining air viscosity and upper surface friction factors.

Using equation (2), this becomes, in terms of geostrophic winds:

(4) $F_P = K_2\ (K_1\ V_G)^2$

Again using equation (1), the retarding force on the floe due to water stress is:

(5) $F_R = -K_3\ V_F^2$ where V_F is the floe speed in the water and K_3 is related to the water viscosity and lower surface friction characteristics. The negative sign indicates that the water stress retards the floe motion.

When ice is moving at a steady speed, the wind force is balanced by the water stress:

(6) $F_P = -F_R$ or, substituting from (4) and (5):

(7) $K_2(K_1\ V_G)^2 = +K_3\ V_F^2$

Solving this equation for the floe speed, V_F, we have:

(8) $V_F = \left[K_1\sqrt{\frac{K_2}{K_3}}\right] V_G = C\ V_G$ since $K_1\sqrt{\frac{K_2}{K_3}}$ is a constant, C.

In other words, the floe speed is related to the geostrophic wind through a simple constant which is likely to be nearly the same for all floes with the same upper and lower surface friction characteristics.

Considerable efforts have been made to determine with greater precision the values of α (the turning angle) and C (the scalar drift coefficient). The results to date support the following values:

C = .008 (under constrained drift conditions, winter and spring with small ice velocities)

C = .01 (under free drift conditions, summer with higher ice velocities)

C = $.008 \leq C \leq .01$ (intermediate drift conditions)

$\alpha = 30\ e^{-1.7\ V_G}$ (α decreases as the geostrophic wind increases)

where e is the base of the system of the natural logarithms ($\sim$2.71828)

Typical values given by this equation are:

$\alpha = 5°$ (to the right) (for moderate wind speeds $\sim$12 kt)

$\alpha = 18°$ (to the right) (for low wind speeds $\sim$5 kt)

3.2.3 Application of Wind-Drift of Ice Equations. The equations for wind-drift of ice given in the previous section are generally correct for drift ice conditions in areas far [100-150 nautical miles (nm)] from the influence of shore or shorefast ice. If the motion of ice under these conditions is to be forecast, site-specific alterations to these rules may need to be employed. Furthermore, although these relationships are generally correct, they represent data with an uncertainty of around $\pm$ 20%. Therefore, when applying these equations, one should not expect the prediction of any one ice event to be any more accurate than $\pm$ 20%.

3.3 CURRENT-DRIFT OF ICE

Oceanic currents in deep arctic and antarctic waters appear to be relatively constant. This contrasts with shallower areas where meteorologically driven currents are possible. Hence, published current charts can be used in these areas. The currents shown are usually on the order of 1 nm/day. Their directions, as shown, are probably correct to within a few degrees. Studies have shown that the current-driven component of ice drift is most accurate when used to predict drift over periods of time on the order of a month or more. On shorter time scales, local fluctuations and deviations of currents become significant.

Current-drift of ice appears to be completely free from shoreline interactions at distances greater than 250 nm from shore. Care must be taken when predicting drift ice behavior based on currents and winds at distances closer than this since it is possible that drift rates may be less.

3.3.1 Short-Term Current-Drift of Ice. The short-term component of ice drift can become very important in regions near shore. In these regions the magnitude of currents can be considerably larger than in areas with deep ocean currents because of meteorological forcing and bathymetric configuration. In these cases, coriolis effects may become significant, tending to pile ice against coastlines to the right of a current or creating a polynya to its left. These currents are difficult to predict. Their behavior depends to a large degree on site-specific interactions which, if understood at all, are cataloged for each general area.

3.4 *INTERNAL FORCES WITHIN THE DRIFT ICE AND THEIR INFLUENCE ON ICE MOVEMENT AND DEFORMATION.*

It has been known for a long time that large forces can develop within drift ice. While the most dramatic demonstration of these forces has been the occasional crushing of ships caught in the ice, the most common indication of large forces within the ice is the occurrence of pressure ridges. Pressure ridges occur when the internal forces within the ice pack exceed the failure strength of the ice. The fact that pressure ridges can be found almost anywhere within the ice demonstrates that such forces are common.

For some time it has been thought that in order to accurately predict drift ice motions, it was necessary to take into account these internal ice forces. This adds a great deal of complication to the forecasting of ice motions since these forces can be transmitted great distances and arise from distant sources. Recently, however, it has been shown that even in close pack conditions, it may not always be necessary to take these forces into account in an explicit way.

Internal forces within the ice are generally even less important when predicting the motion of the drift ice edge than when predicting movements within the ice. As a result of this, ice analysts generally do not take internal forces into account in a rigorous way when predicting pack advances and retreats.

The treatment just given for the motion of ice under the influence of winds and currents is essentially an outline of the basic concepts. In Chapter IV two approaches are described that deal with the problem of predicting sea ice motions.